

Warm Atlantic surface water inflow to the Nordic seas 34–10 calibrated ka B.P.

Tine L. Rasmussen^{1,2} and Erik Thomsen³

Received 14 March 2007; revised 11 September 2007; accepted 10 October 2007; published 15 January 2008.

[1] A number of short-lasting warm periods (interstadials) interrupted the otherwise cold climate of the last glacial period. These events are supposedly linked to the inflow of the warm Atlantic surface water to the Nordic seas. However, previous investigations of planktonic foraminifera from the Nordic seas have not been able to resolve any significant difference between the interstadials and intervening cold stadials, as the faunas are continuously dominated by the polar species *Neogloboquadrina pachyderma* s. Here we examine the planktonic foraminifera assemblages from a high-resolution core, LINK17, taken at 1500 m water depth off northern Scotland below the warmest part of the inflowing Atlantic water. The core comprises the time period 34–10 calibrated ka B.P., the coldest period of the last glaciation and the deglaciation. The results reveal a hitherto unknown faunistic variability indicating significant fluctuations in both surface water inflow and in summer sea surface temperatures. During the interstadials, relatively warm Atlantic surface water (4–7°C) flowed north into the eastern Norwegian Sea. During the stadials and Heinrich events the surface inflow stopped and the temperatures in the study area dropped to <2°C. The Last Glacial Maximum was nearly as warm as the interstadials, but the inflow was much more unstable. The data reveal two previously unrecognized warming events each lasting more than 1600 years and preceding Heinrich events HE3 and HE2, respectively. By destabilizing the ice sheets on the shelves the warmings may have played a crucial role for the development of Heinrich events HE2 and HE3.

Citation: Rasmussen, T. L., and E. Thomsen (2008), Warm Atlantic surface water inflow to the Nordic seas 34–10 calibrated ka B.P., *Paleoceanography*, 23, PA1201, doi:10.1029/2007PA001453.

1. Introduction

[2] Ocean currents are of great importance to the climate of land areas surrounding the North Atlantic Ocean and Nordic seas (Figure 1). The warm Atlantic surface water flowing into the eastern part of the Nordic seas is responsible for the relatively mild winter climate of northern Europe. In the Nordic seas the warm surface water is cooled and convected to form cold deep water masses, which overflow the Greenland-Scotland Ridge and return to the North Atlantic Ocean.

[3] Previous results from core ENAM93-21 north of the Faeroe Islands (Figure 1) indicate that there was outflow of deep water from the Nordic seas during the warm interstadials of marine isotope stage (MIS) 3 and MIS 2 [Rasmussen *et al.*, 1996a, 1996b]. However, the core gave no clear evidence for a compensating inflow of warmer Atlantic surface water, as the interstadials were totally dominated by the polar planktonic foraminifera *Neogloboquadrina pachyderma* s [Rasmussen *et al.*, 1996a]. The assemblages of ENAM93-21 are typical for the interstadials of the Nordic

seas, where subpolar species generally make up less than 5% of the faunas and the interstadials appear almost as cold as the stadials [e.g., Fronval *et al.*, 1995; Rasmussen *et al.*, 1996a, 1996b; Voelker *et al.*, 1998; Weinelt *et al.*, 2003]. This is surprising as the Greenland ice core records indicate a considerable climate shift between stadials and interstadials (Dansgaard-Oeschger events) [Johnsen *et al.*, 1992; Dansgaard *et al.*, 1993]. It is also in strong contrast to the conditions in the North Atlantic south of the Greenland-Scotland Ridge, where subpolar planktonic foraminifera often constitute more than 50–70% of the interstadial faunas indicating a significant interstadial surface warming [Bond *et al.*, 1993; Oppo and Lehman, 1995; van Kreveld *et al.*, 2000; Weinelt *et al.*, 2003; Rasmussen *et al.*, 2003a]. Given the presence of convection and the outflow of cold deep water [Rasmussen *et al.*, 1996a] a balancing inflow of warmer surface water would be expected. The aim of the present study is to examine and test this hypothesis.

[4] Today the inflow of Atlantic surface water to the Nordic Seas occurs mainly in the area around the Faeroe Islands with the warmest inflow concentrated east of the islands along the British Isles (Figure 1). The focus of the present study is on a high-resolution core, LINK17, retrieved from the eastern side of the Faeroe-Shetland Channel [Nielsen *et al.*, 2000] almost exactly below the path of the warmest inflow. The core contains the time period 34–10 calibrated ka B.P. This period has been intensively studied in several cores from both west, north, and south of the islands (ENAM93-21, ENAM33, and DAPC-02) thus

¹Department of Geology, University of Tromsø, Tromsø, Norway.

²Temporarily at Woods Hole Oceanographic Institute, Woods Hole, Massachusetts, USA.

³Department of Earth Sciences, University of Aarhus, Aarhus, Denmark.

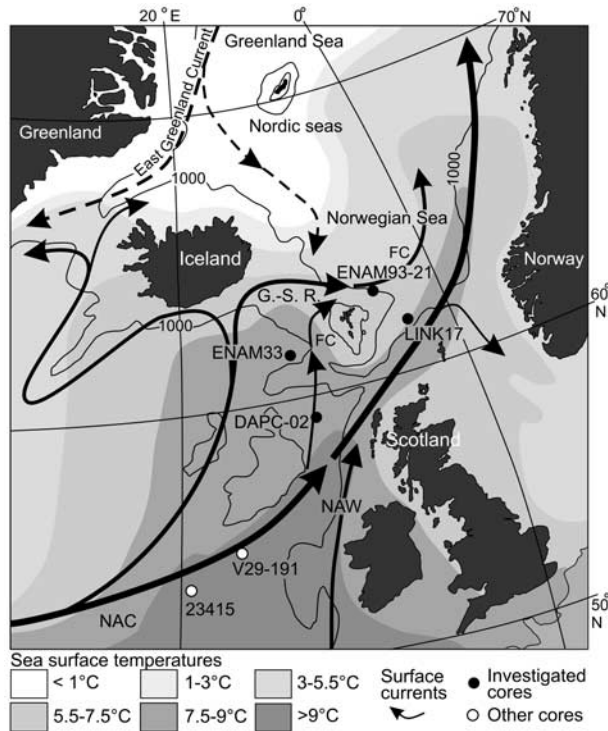


Figure 1. Map of northeastern North Atlantic Ocean and southeastern Nordic seas showing present surface currents and surface temperatures for the month of March (modified from Orvik and N  ler [2002]). Location of cores is indicated. Abbreviation is G.-S.R., Greenland-Scotland Ridge.

allowing a detailed investigation of the whole inflow area. The investigated time period comprises the late Marine Isotope Stage (MIS) 3 and MIS 2. It includes Heinrich events HE3, HE2, and HE1; the warm interstadials IS4, IS3, and IS2; and the Last Glacial Maximum (LGM) [e.g., Heinrich, 1988; Johnsen *et al.*, 1992; Dansgaard *et al.*, 1993; Bond *et al.*, 1993].

[5] A potentially important reason for the lack of evidence of warm surface water in the Nordic seas during the glacial period is poor planktonic foraminifera assemblages recovered from this period. As noted above, they are generally dominated by *N. pachyderma* s. The low amount of faunistic variation may be enhanced by the large mesh size used in many investigations. Most studies aiming at calculating sea surface temperatures (SST) by transfer functions are based on the $>150 \mu\text{m}$ fraction. This is unfortunate as planktonic foraminifera in cold Arctic and Polar waters are small; especially those associated with high productivity zones in the vicinity of ocean fronts or in the productive Arctic surface water in the central Nordic seas [e.g., B   and Tolderlund, 1971; Johannessen *et al.*, 1994; Carstens *et al.*, 1997]. One of the most important species living in Atlantic surface water at high northern latitudes is *Turborotalita quinqueloba* [B   and Tolderlund, 1971; Carstens *et al.*, 1997]. In investigations based on mesh sizes >125 or $>150 \mu\text{m}$ the signal of *T. quinqueloba* is

often lost [Pflaumann *et al.*, 2003], except when conditions are fully interglacial [Kandiano and Bauch, 2002]. However, much of the information can be retained using $100 \mu\text{m}$ mesh-size sieves [Rasmussen *et al.*, 1996a; Kandiano and Bauch, 2002]. The records used in the present study are all based on the $>100 \mu\text{m}$ size fraction.

2. Study Area

[6] The Faeroe-Shetland area is the primary gateway for the inflow of warmer Atlantic surface water to the Nordic seas (Figure 1). At present the inflow is divided into two current branches, one on each side of the Faeroe Islands (Figure 1). The eastern branch, the North Atlantic water, flows northward on the eastern side of the Faeroe-Shetland Channel along the British Isles [van Aken, 1988; Sherwin *et al.*, 1999; Hansen and   sterhus, 2000; Orvik and N  ler, 2002]. The western branch, the North Atlantic Current continuing as the Faeroe Current, passes west of the Faeroe Islands. The Faeroe Current is affected by colder water from the polar East Icelandic Current, which is a branch of the polar East Greenland Current. The Iceland-Faeroe Front separates the cold surface water from the warmer Atlantic water northeast of the Faeroe Islands. The North Atlantic water is warmer and more saline than the Faeroe Current.

3. Materials

[7] The focal core of this study, LINK17, is located below the eastern branch of the Atlantic water flowing into the Nordic seas. The 985 cm long piston core was retrieved from a water depth of 1500 m during the August 2000 R/V DANA cruise (Figure 1). The site is situated about 35 km south of a submarine fan, the North Sea fan, at the mouth of the Norwegian Channel. The fan is formed mostly by Quaternary glaciogenic debris-flow deposits intercalated with interglacial hemipelagic sediments [e.g., Sejrup *et al.*, 1996; King *et al.*, 1996]. The core consists of homogenous, hemipelagic mud sometimes containing dropstones. Previous work on the core comprises geochemical analysis of a single sample of tephra grains taken from an ash layer at 633 cm and dated to 23,850 ^{14}C years [Wasteg  rd *et al.*, 2006].

[8] We compare results from LINK17 to those from ENAM93-21, ENAM33, and DAPC-02. Core ENAM93-21 is located in a colder area 250 km west of LINK17 in the pathway of the Faeroe Current [Rasmussen *et al.*, 1996a, 1996b]. Core ENAM33 is located 525 km southwest of LINK17 and south of the Greenland-Scotland Ridge [Rasmussen *et al.*, 2003a]. Core DAPC-02 is from the northern Rockall Plateau 475 km southwest of LINK17 [Rasmussen *et al.*, 2002] (Figure 1).

4. Methods

[9] The core was sampled every 5 cm and in some intervals every 2 cm. The samples (approximately 6 cm^3 each) were dried, weighed, and sieved over 63 and $100 \mu\text{m}$ sieves. Grain sizes are presented as the weight percentages of the fraction $>100 \mu\text{m}$ and the fraction 63– $100 \mu\text{m}$, respectively. At least 300 planktonic and 300 benthic

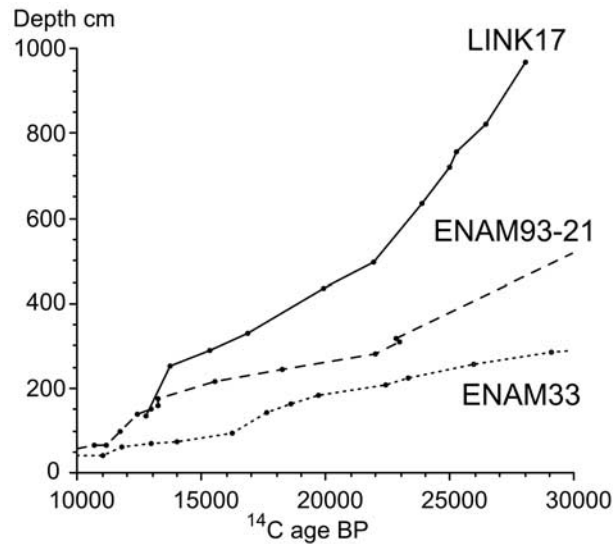


Figure 2. Age-depth plot for cores LINK17, ENAM93-21, and ENAM33.

foraminifera were picked from the $>100\ \mu\text{m}$ residue from each sample using a picking tray. The concentrations of planktonic and benthic foraminifera were calculated as the number per gram dry weight sediment. Tephra grains were also counted in the $>100\ \mu\text{m}$ fraction and concentrations calculated. The samples were subsequently dry sieved on $150\ \mu\text{m}$ sieves and >300 mineral grains representing ice-rafted detritus (IRD) was counted and the concentration per gram dry weight sediment calculated. The $\delta^{18}\text{O}$ values were measured on specimens of the planktonic foraminifera *N. pachyderma* s at Woods Hole Oceanographic Institution using standard procedures [Ostermann and Curry, 2000]. Accelerator mass spectrometry (AMS) ^{14}C dates were measured at the AMS ^{14}C Laboratory, University of Aarhus on *N. pachyderma* s with one exception (Figure 2, Table 1). The dates were corrected for a reservoir effect of -400 years and calibrated using the calibration curve of Fairbanks *et al.* [2005], version Fairbanks0107 (Table 1).

[10] Piston core ENAM93-21 is from 1020 m water depth (Figure 1). Results on foraminifera, IRD, oxygen isotopes,

and AMS- ^{14}C dates from this core were published by Rasmussen *et al.* [1996a, 1996b, 1996c, 1998, 2003b]. For this study, the upper part comprising the last 30,000 years has been resampled at 2.5 cm intervals to increase the time resolution and ensure that all short-lasting events were captured. The same methods as for LINK17 were used [see also Rasmussen *et al.*, 1996a, 1996b]. Core ENAM33 is from 1217 m water depth. The foraminifera, IRD, oxygen isotope data, and AMS- ^{14}C dates were published by Kuijpers *et al.* [1998] and Rasmussen *et al.* [2003b]. Only the upper part of the core, younger than 34.5 calibrated ka B.P., is included in this study (Figure 2). Piston core DAPC-02 is from 1709 m water depth. Oxygen isotope data published by Knutz *et al.* [2002, 2007] and foraminifera data from Rasmussen *et al.* [2002] are included in the discussion for comparison of events leading to the Heinrich event HE1. For this study, IRD was counted in the $>100\ \mu\text{m}$ fraction.

5. Results

[11] The stratigraphic intervals are identified by correlation with the Greenland ice cores and by combining the AMS- ^{14}C ages, the oxygen isotope record, the grain size distribution, IRD content, and the faunistic records (Figures 2, 3, and 4). The AMS- ^{14}C dates show that the time period 34–10 calibrated ka B.P. is present in very high resolution (Figures 2, 3, and 4) with an average sedimentation rate of $>50\ \text{cm/ka}$, meaning that a 1 cm thick sample represents approximately 17 years. Features on a timescale of 50 years or less should be clearly resolvable.

[12] LINK17 can be correlated to the Greenland North-GRIP and GRIP ice cores with great precision (Figure 4). A tephra layer 633 cm down core in LINK17 has been identified as the Fugloyarbanki Tephra, also called the Faeroe Marine Ash Zone (FMAZ) II [Wastegård *et al.*, 2006]. This distinct ash layer has been geochemically identified in several cores from the southern Nordic seas including ENAM93-21 and ENAM33 [Rasmussen *et al.*, 2003a; Wastegård *et al.*, 2006]. In LINK17 the layer is dated to 28,600 calibrated years B.P., which is somewhat older than the previously obtained dates, which average at around 27,600 calibrated years B.P. (Figures 4, 5, and 6). The Fugloyarbanki Tephra has recently been identified in the NorthGRIP record also by geochemical analysis [Davies *et al.*, 2007; Wastegård *et al.*, 2006]. The tephra is situated

Table 1. AMS ^{14}C Dates and Calibrated Dates for Core LINK17

Depth, cm	AMS ^{14}C , –400 years ^a	Laboratory Code	Species	Calibrated Age (Fairbanks 0107)
135	12,760 \pm 150	AAR-7236	<i>N. pachyderma</i> s	14,860 \pm 210
250	13,720 \pm 120	AAR-7237	<i>N. pachyderma</i> s	15,970 \pm 180
290	15,330 \pm 110	AAR-8137	<i>N. pachyderma</i> s	18,570 \pm 120
331	16,880 \pm 140	AAR-7238	<i>N. pachyderma</i> s	20,050 \pm 170
436.5	19,930 \pm 190	AAR-7239	Bivalve	23,810 \pm 190
496.5	21,900 \pm 220	AAR-8138	<i>N. pachyderma</i> s	26,350 \pm 280
633	23,850 \pm 350	AAR-7240	<i>N. pachyderma</i> s	28,566 \pm 420
718.5	24,950 \pm 300	AAR-8139	<i>N. pachyderma</i> s	29,950 \pm 450
754	25,250 \pm 350	AAR-8140	<i>N. pachyderma</i> s	30,360 \pm 470
818.5	26,450 \pm 350	AAR-8141	<i>N. pachyderma</i> s	31,720 \pm 400
965	28,050 \pm 450	AAR-8142	<i>N. pachyderma</i> s	33,420 \pm 500

^aReservoir effect.

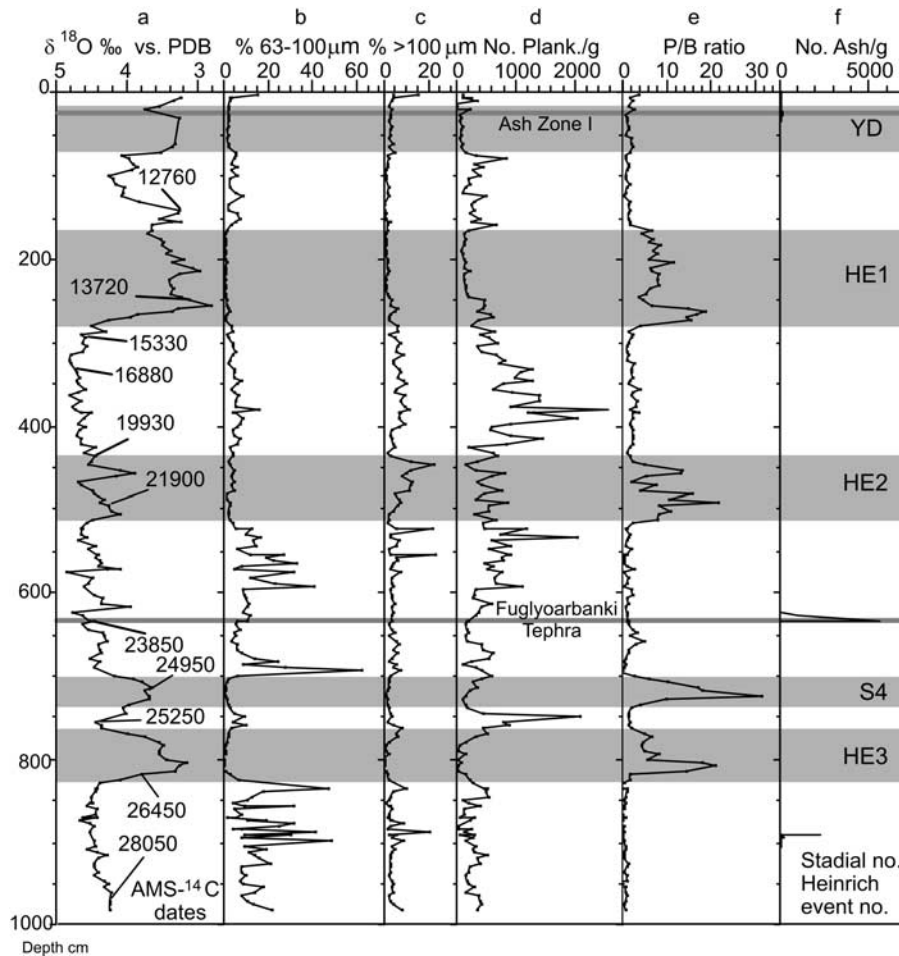


Figure 3. Records plotted versus depth down core of core LINK17. (a) Planktic $\delta^{18}\text{O}$ values measured on *N. pachyderma* s with AMS ^{14}C dates marked. (b) Weight percent of grain size fraction 63–100 μm . (c) Weight percent of grain size fraction >100 μm . (d) Number of planktonic foraminifera per gram dry weight sediment. (e) Planktic/benthic ratio (number of planktic foraminifera/number of benthic foraminifera per gram dry weight sediment). (f) Number of tephra grains per gram dry weight sediment. Position of Fugloyarbanki Tephra and Younger Dryas Ash Zone I is marked. Abbreviations are YD, Younger Dryas; HE, Heinrich events; S, stadial.

above the warmest phase of interstadial (IS) 3 [Rasmussen *et al.*, 2003b; Davies *et al.*, 2007] (Figure 4). With this tie point it is easy to identify IS4, IS3, IS2, and IS1 (the Bølling and Allerød interstadials) in the LINK17 core (Figures 3 and 4). As in the ice core, the interstadials are short lasting and begin and end abruptly (Figure 4). The only exception is the boundary between HE1 and the Bølling interstadial, which is stepwise (Figure 4). We have placed the boundary at the simultaneous abrupt shifts in the concentration of planktonic foraminifera and planktic/benthic (P/B) ratio (Figure 3). Apparently, the LINK17 core does not reach into interstadial IS5, which thus is older than 33,400 calibrated years B.P. Interstadial IS4 is centered at 30,400 calibrated years B.P., IS3 dates shortly after 30,000 calibrated years B.P., while IS2 begins around 24,000 calibrated years B.P. (Figure 4).

[13] Interstadial events are characterized by low percentages of *N. pachyderma* s and high percentages of the subpolar planktonic species, varying between 40 and 70%.

T. quinqueloba is the most abundant species (Figure 4), but *N. pachyderma* d, *Globigerina bulloides*, *Globigerinita glutinata*, *G. uvula*, *Globorotalia inflata*, and *Orbulina universa* [Bé and Tolderlund, 1971] are also present. *Neogloboquadrina pachyderma* d is relatively rare except for the Bølling and Allerød interstadials and the Holocene. The interstadial intervals correlate with generally high $\delta^{18}\text{O}$ values, low concentrations of IRD, high concentrations of planktonic foraminifera, and low P/B ratios (Figure 3). The top of the core is marked by a thin layer of probably interglacial Holocene deposits characterized by low $\delta^{18}\text{O}$ values and a high relative abundance of the subpolar foraminifera (up to 60% of *N. pachyderma* d) (Figure 4). *G. inflata* and *O. universa* are present in low numbers.

[14] Heinrich events and stadials are identified on the basis of low- $\delta^{18}\text{O}$ values, high IRD content, a dominance of *N. pachyderma* s, and a relatively low concentration of planktonic foraminifera. These features are typical for Heinrich events in North Atlantic records and easily recog-

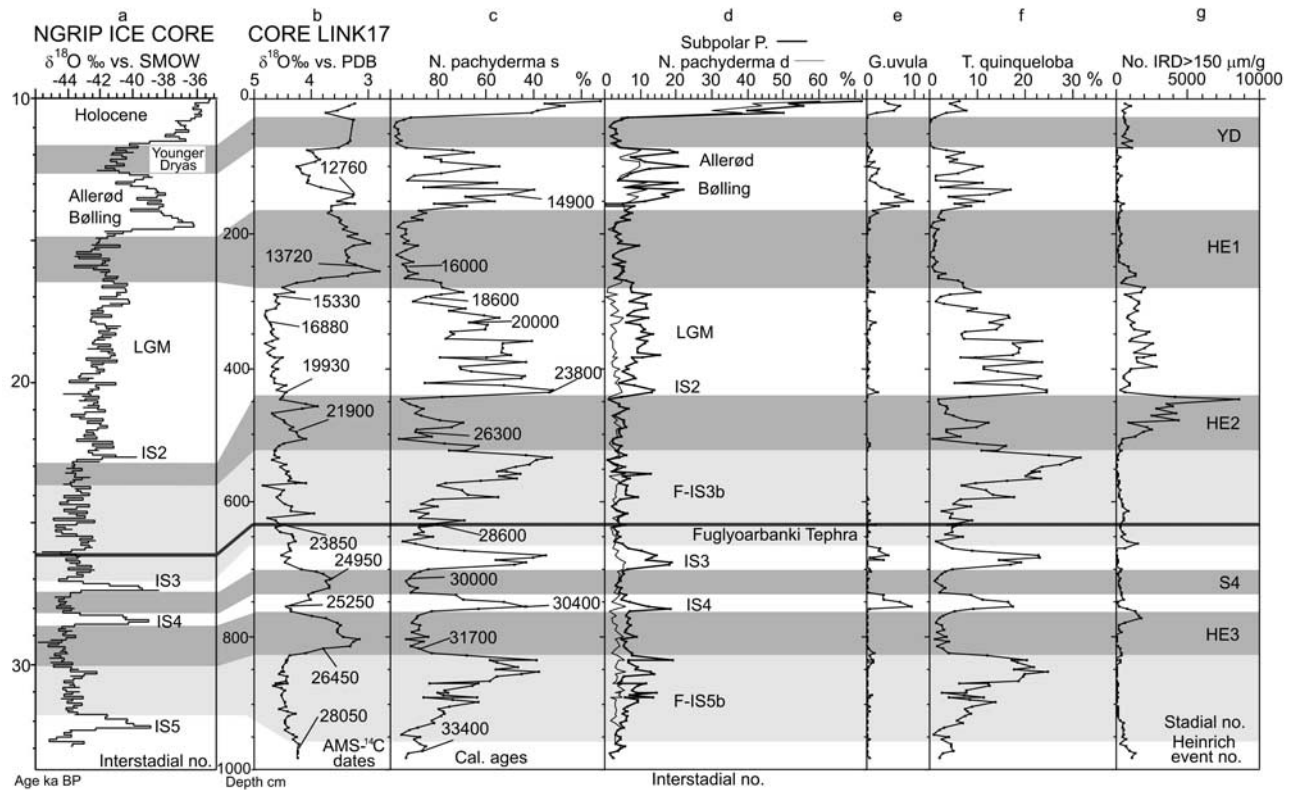


Figure 4. Planktic $\delta^{18}\text{O}$ values and variation in relative abundance of foraminifera species for core LINK17 correlated with the $\delta^{18}\text{O}$ record of the NorthGRIP ice core. (a) NorthGRIP ice core $\delta^{18}\text{O}$ record (data from North Greenland Ice Core Project members, 2004). (b) Planktic $\delta^{18}\text{O}$ values measured on *N. pachyderma* s with AMS ^{14}C dates marked. (c) Percent of *N. pachyderma* s with calibrated ages marked. (d) Cumulative percent of the subpolar species *N. pachyderma* d, *G. bulloides*, *G. glutinata*, *G. inflata* and *O. universa*. (e) Percent of *G. uvula*, and (f) percent of *T. quinqueloba*. (g) Number of IRD grains $>150\ \mu\text{m}$ per gram dry weight sediment. Position of Fugloyarbanki Tephra is marked on ice core according to results by Davies *et al.* [2007]. Abbreviations are as in Figure 3 but also with LGM, Last Glacial Maximum; IS, interstadial.

nized [e.g., Heinrich *et al.*, 1988; Bond *et al.*, 1993] (Figures 3 and 4). The high P/B ratio is typical for stadial and Heinrich events around the Faeroe Islands [see Rasmussen *et al.*, 1996a] (Figure 3). Heinrich event HE3 includes a date of 31,700 calibrated years B.P., while HE2 and HE1 are bracketed by dates of 26,300–23,800 calibrated years B.P. and 18,600–14,900 calibrated years B.P., respectively (Figures 3 and 4, Table 1). A sample within the HE1 event is dated to 16,000 calibrated years B.P. The ages agree well with other dates of HE3, HE2, and HE1 obtained from the Nordic seas and North Atlantic [e.g., Bond *et al.*, 1993; Fronval *et al.*, 1995; Voelker *et al.*, 1998; van Kreveld *et al.*, 2000].

[15] The LGM is distinguished by high $\delta^{18}\text{O}$ values, high abundances of planktonic foraminifera, and a high proportion of IRD and grains in the size fraction $>100\ \mu\text{m}$ (Figure 3). The interval is characterized by strong fluctuations in the relative abundance of *N. pachyderma* s (Figure 4). The LGM spans from 23,000 to around 19,000 calibrated years B.P.

[16] In LINK17, IS5 and IS3 are followed by long intervals with high $\delta^{18}\text{O}$ values, high concentrations of planktonic foraminifera, low-P/B ratios, decreasing concentrations of IRD, and high sedimentation rates (Figures 2, 3,

and 4). In the planktonic faunas, *T. quinqueloba* increases in relative abundance reaching a maximum of about 60% of the fauna, whereas *N. pachyderma* s decreases. In most parameters the intervals resemble the interstadials. However, comparison with the NorthGRIP ice core shows that the two intervals here correlate with relatively cold phases (Figure 4). Since the two events are only faintly visible in the ice core, and apparently have a limited geographical distribution, while following the preceding interstadials IS5 and IS3, we designate them F-IS5b and F-IS3b (Faeroe-interstadial) from the area where they are well developed. F-IS5b lasts >1600 years, while the F-IS3b lasts about 2000 years. The two intervals terminate with HE3 and HE2, respectively (Figure 4).

6. Discussion

6.1. Paleooceanographic Interpretation of LINK17

[17] The large quantity of subpolar species and the almost total absence of IRD in the interstadials of LINK17 indicate inflow of Atlantic water to the Nordic seas and warm conditions during these periods (Figure 4, Table 2). High

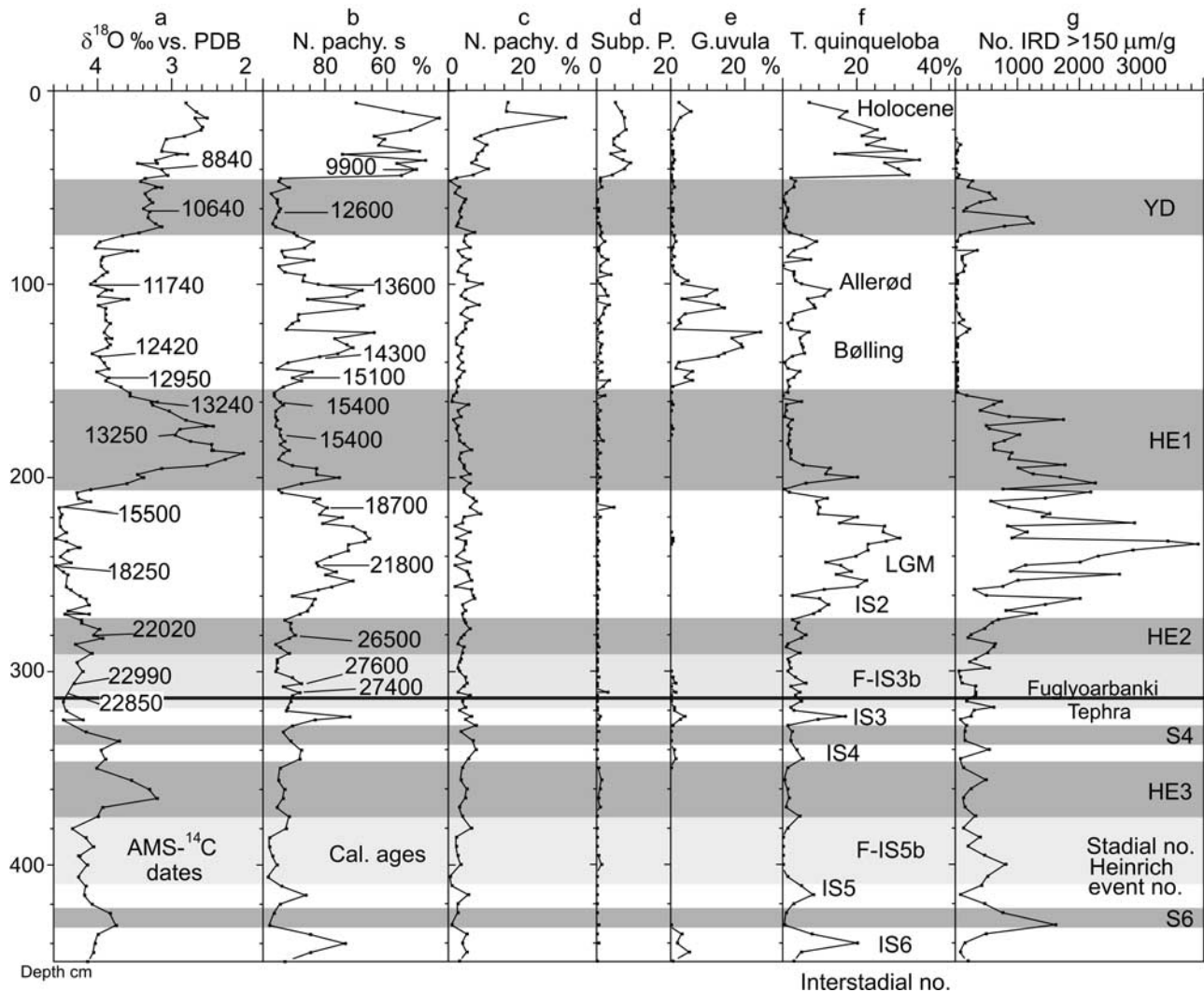


Figure 5. Planktic $\delta^{18}\text{O}$ values and variation in relative abundance of foraminifera species and IRD for core ENAM93-21. (a) Planktic $\delta^{18}\text{O}$ values measured on *N. pachyderma* s with AMS ^{14}C dates marked. (b) Percent of *N. pachyderma* s with calibrated ages marked. (c) Percent of *N. pachyderma* d. (d) Cumulative percent of subpolar species *G. glutinata* and *G. bulloides*. (e) Percent of *G. uvula*, and (f) percent of *T. quinqueloba*. (g) Number of IRD grains $>150\ \mu\text{m}$ per gram dry weight sediment. Position of Fugloyarbanki Tephra is marked. Abbreviations as in Figures 3 and 4.

percentages of *T. quinqueloba* are typically found in Arctic and Atlantic surface water, where the nutrient content is high [Johannessen et al., 1994; Carstens et al., 1997] (Table 2). Today the species is abundant on the western Svalbard margin in cooled Atlantic water with high productivity [Johannessen et al., 1994; Carstens et al., 1997] and summer temperatures varying from 4 to 7°C (Table 2). *G. uvula* is an opportunist species, abundant at 30–55°S in the South Atlantic and in upwelling areas [Boltovskoy et al., 1996; Kemle-von Mücke and Hemleben, 1999].

[18] At the beginning of the two interstadials F-IS5b and F-IS3b the surface conditions at LINK17 were very cold as indicated by high percentages of *N. pachyderma* s and relatively high IRD concentrations (Figure 4). The inflow of Atlantic water was probably reduced. The gradual

increase in the percentage of *T. quinqueloba* during both F-IS5b and F-IS3b and the concomitant decrease in IRD indicate a gradual warming, which reached a maximum just prior to HE3 and HE2, respectively (Figure 4). Surface water conditions at LINK17 were probably not very different from the interstadial conditions with summer temperatures varying between 4 and 7°C (Table 2). Two sudden depletions indicate that the $\delta^{18}\text{O}$ values varied more during the F-IS3b event than during the F-IS5b event (Figure 4). Both depletions correlate with colder surface water as demonstrated by the corresponding increases in the relative abundance of *N. pachyderma* s.

[19] The total dominance of *N. pachyderma* s and the high concentration of IRD during stadials and Heinrich events (Figure 4) indicate that surface conditions must have been

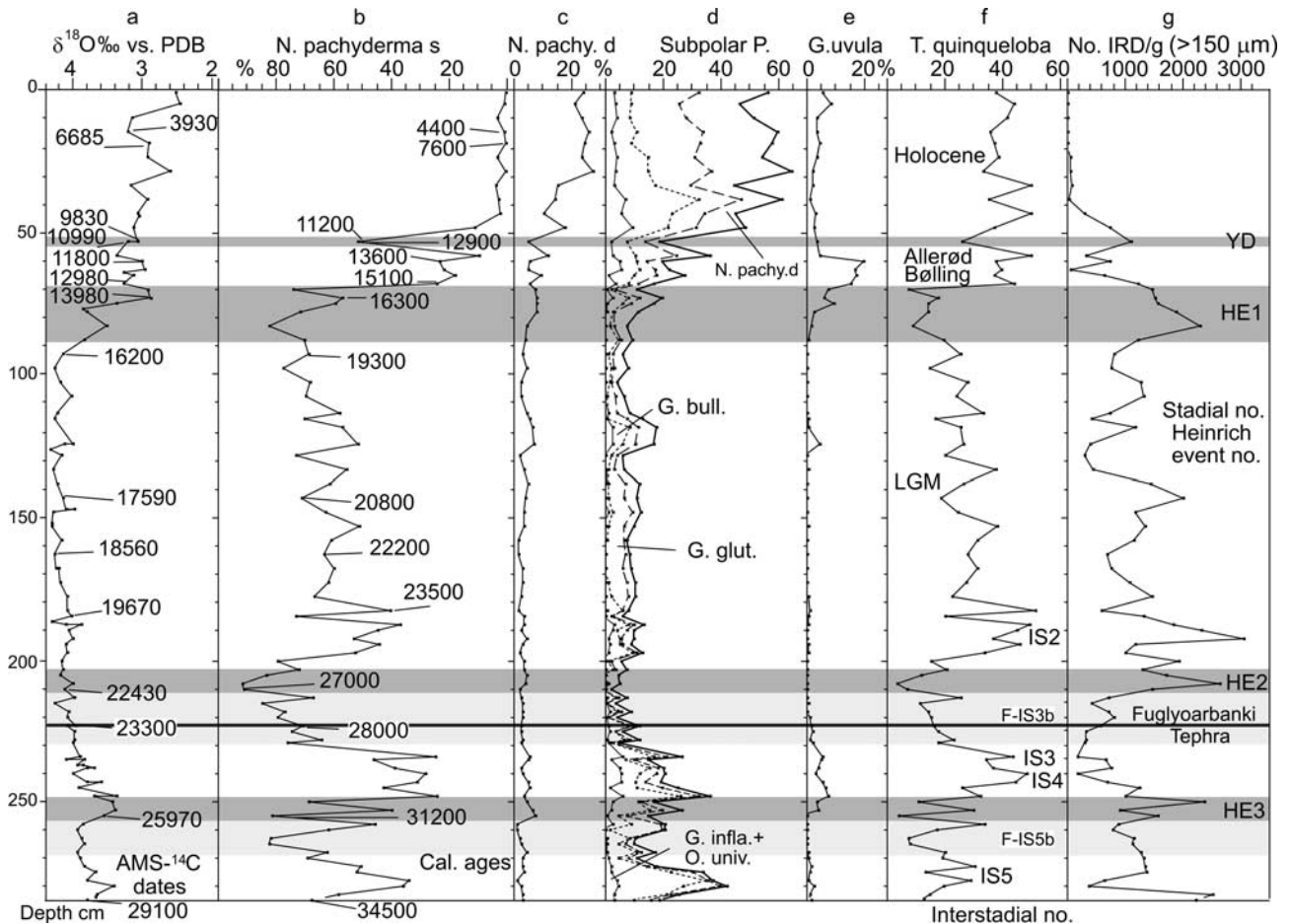


Figure 6. Planktic $\delta^{18}\text{O}$ values and variation in relative abundance of foraminifera species and IRD for core ENAM33. (a) Planktic $\delta^{18}\text{O}$ values measured on *N. pachyderma* s with AMS ^{14}C dates marked. (b) Percent of *N. pachyderma* s with calibrated ages marked. (c) Percent of *N. pachyderma* d. (d) Cumulative percent of subpolar species *G. inflata* and *O. universa*, *G. bulloides*, *G. glutinata*, and *N. pachyderma* d. (e) Percent of *G. uvula*, and (f) percent of *T. quinqueloba*. (g) Number of IRD grains $>150\ \mu\text{m}$ per gram dry weight sediment. Position of Fugloyarbanki Tephra is marked. Abbreviations as in Figures 3 and 4.

Table 2. Summary of the Environmental Preferences of the Planktonic Foraminifera Species

Species	Environmental Conditions/Water Mass	References
<i>N. pachyderma</i> s	Polar conditions, 40–90% in Arctic surface water (SST = 3–5°C) and >95% in Polar surface water (SST < 2°C)	<i>Bé and Tolderlund, 1971; Johannessen et al., 1994; Pflaumann et al., 1996.</i>
<i>N. pachyderma</i> d, <i>G. bulloides</i> , and <i>G. glutinata</i>	Subpolar conditions, 20–70% in warm Atlantic surface water in the Nordic seas (SST = 7–14°C)	<i>Bé and Tolderlund, 1971; Johannessen et al., 1994; Carstens et al., 1997.</i>
<i>T. quinqueloba</i>	Subpolar and arctic conditions, 40–80% in productive arctic water and at the Arctic Front (SST = 4–7°C) and 10–20% at the Polar Front	<i>Bé and Tolderlund, 1971; Johannessen et al., 1994; Carstens et al., 1997.</i>
<i>G. uvula</i>	Polar front and upwelling areas, cold productive surface water, opportunist species, wide SST range of ~3–15°C	<i>Parker, 1962; Ottens and Nederbragt, 1992; Boltovskoy et al., 1996; Kemle-von Mücke and Hemleben, 1999</i>
<i>G. inflata</i> and <i>O. universa</i>	Transitional and subtropical conditions SST > 12–15°C	<i>Bé and Tolderlund, 1971</i>

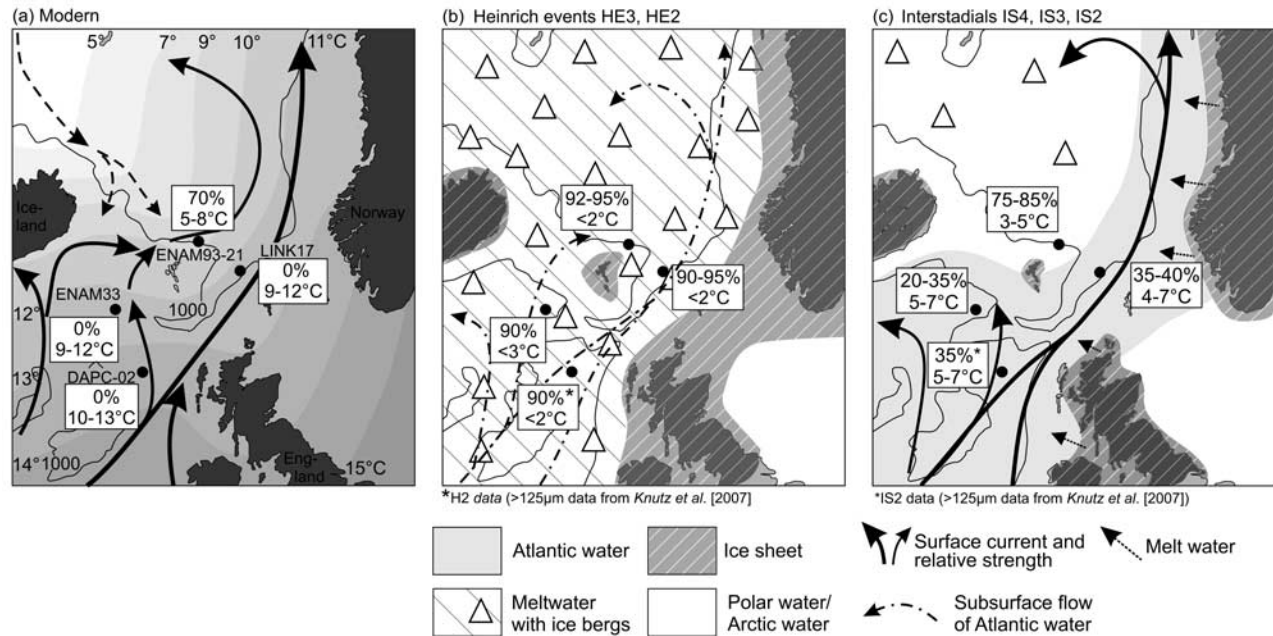


Figure 7. (a) Map showing present summer sea surface temperatures and current systems in the northeastern Atlantic and southeastern Nordic seas. (b and c) Paleocceanographic reconstructions for the same area as in Figure 7a showing sea surface temperatures, surface and near surface current systems, and the distribution of major water masses for interstadials (IS4, IS3, and IS2) and Heinrich events and stadials (HE3, HE2, and S4), respectively. The inserts show the percentage of *N. pachyderma* s (upper row) and the estimated summer sea surface temperatures (lower row) for each of the cores LINK17, ENAM93-21, ENAM33, and DAPC-02. The temperature estimates are based on the composition of planktonic faunas and the planktic $\delta^{18}\text{O}$ values.

cold and polar with drifting icebergs. The ice cover was probably extensive and summer SST below 2°C (Table 2). The low $\delta^{18}\text{O}$ values are usually taken as a signal of meltwater and the surface water was undoubtedly stratified [e.g., Bond *et al.*, 1993]. The transitions between stadial-interstadial events are abrupt with sudden warmings and coolings. The only exceptions are the gradual surface warming at the beginning of the Bølling interstadial, and a small warming event, indicated by an increase in *T. quinqueloba*, that occurs at the start of the HE1 (Figure 4).

[20] The LGM interval, 23,000–19,000 calibrated years B.P., resembles the interstadials more than it resembles the stadials except for the high, but fluctuating concentration of IRD (Figure 4). The proportion of subpolar species was relatively high and *T. quinqueloba* was abundant constituting up to 50% of the planktonic foraminifera (Figure 4). The faunas indicate Arctic conditions with inflow of Atlantic surface water. The maximum sea surface temperatures during the LGM were probably almost the same as during the interstadials.

6.2. Flow of Atlantic Surface Water into the Nordic Seas During Late MIS 3 and MIS 2

[21] Core LINK17 demonstrates unambiguously that during interstadials IS4, IS3, and IS2 Atlantic surface water flowed into the Nordic seas (Figure 7c). In order to better constrain the inflow area and temporal change in the inflow

we compare the result of LINK17 with the records from the nearby cores ENAM93-21, ENAM33, and DAPC-02. ENAM93-21 is located in a colder area north of the Faeroe Islands, while ENAM33 and DAPC-02 are from the warmer North Atlantic south of the islands. In the discussion we concentrate on the warm periods, namely, the interstadials and the LGM. The records of stadials and Heinrich events are very similar in the examined cores indicating cold and stratified conditions over the entire study area with no surface inflow of Atlantic water to the Nordic seas (Figure 7b).

6.2.1. Interstadials IS4, IS3, and IS2

[22] The closer sampling of ENAM93-21 confirms that the interstadials at this site were continuously dominated by *N. pachyderma* s (Figure 5) (see also Rasmussen *et al.* [1996a]). The sea surface was constantly cold and the conditions must have been very different from those at LINK17. Considering the abundance of *N. pachyderma* s and the low concentration of *T. quinqueloba*, it is probable that the water at ENAM93-21 site was cooled Atlantic surface water flowing from the southwest. The site was undoubtedly also affected by the cold East Greenland Current (Figure 7c).

[23] The conditions at ENAM33, located at 61°N south of the Greenland-Scotland Ridge, were fairly similar to those at LINK17. The abundance of *G. bulloides*, *G. inflata*, and other subpolar species varies between approximately 60 and 80%, as compared to 40–70% in LINK17, and less than

15% in ENAM93-21 (Figures 4, 5, and 6). Summer surface temperatures were probably 4–7°C and nearly the same as at LINK17. The IRD concentration is generally low. Combined, the three cores demonstrate that during interstadials the path of the inflowing warm water was narrow, limited to the eastern part of the Faeroe-Shetland Channel. However, the weak SST gradient between ENAM33 and LINK17 suggests that the inflow of North Atlantic water to the Nordic seas was relatively strong. This is in agreement with the suggestion of Weinelt *et al.* [2003].

[24] The pattern of the interstadials in the records of LINK17, in particular the distribution of subpolar planktonic foraminifera, is very similar to the pattern shown by the $\delta^{18}\text{O}$ records of the Greenland ice cores (Figure 4). The graphs indicate that the transition from cold stadial to warm interstadial conditions was extremely abrupt taking 10–30 years or less. The period with maximum warmth and a strong inflow of Atlantic water lasted only between 50 and 100 years, while the coolings took from 100–200 years (Figure 4).

6.2.2. Interstadials F-IS5b and F-IS3b

[25] The normal interstadials from late MIS 3 and MIS 2 are brief (<~350 years) and abrupt. Furthermore, they are widely, possibly globally, distributed. This is in contrast to interstadials F-IS5b and F-IS3b, which are long lasting (1600–2000 years) with gradual warmings and abrupt coolings and, apparently, a limited geographical distribution.

[26] The most distinctive feature of the two events in LINK17 is the increasing relative abundance of *T. quinqueloba* and other subpolar species. In the other investigated cores, this increase is much less pronounced (Figures 5 and 6). In ENAM33, only the peak abundances just prior to HE3 and HE2, respectively, are apparent, indicating that this site laid outside of the main inflow of Atlantic water to the Nordic seas. In ENAM93-21, only F-IS5b is discernible. The IRD content of the interval is relatively high and variable and conditions were probably continuously cold and polar (Figure 5, Table 2). It is interesting to note that although the sea surface temperatures in the eastern Norwegian Sea at the beginning of both F-IS5b and F-IS3b were low, the $\delta^{18}\text{O}$ values were high (Figure 4). This indicates that the water column was homogeneous as during interstadials and not stratified as during stadials, and it is likely that convection occurred (Figures 4, 5, and 6) [cf. Rasmussen *et al.*, 1996a].

[27] The gradual warmings that affected both F-IS5b and F-IS3b were probably restricted to the eastern part along the British and Fennoscandian ice sheets. We suggest that as the temperatures rose and Atlantic inflow increased, the ice shelves, possibly in conjunction with a sea level rise, became more unstable. Eventually, the ice shelves collapsed and the ice and meltwater stopped the convection. The transition from F-IS5b interstadial to Heinrich event HE3 and from F-IS3b to HE2 took apparently less than 20 years (Figure 4). According to Hulbe *et al.* [2004], a climatic warming could cause rapid breakdown of marine ice shelves and account for the increase in iceberg production during Heinrich events. As meltwater flowed out of the Nordic seas at the surface, compensating Atlantic water flowed northward below the meltwater layer warming the intermediate

water layer in the Nordic seas [Rasmussen and Thomsen, 2004] (Figure 7b). Rasmussen and Thomsen [2004] also suggest that the rapid abrupt warmings at the end of the stadials were caused by a rapid surfacing of this relatively warm water mass.

6.2.3. Atlantic Inflow and the Instability of the LGM

[28] In the Nordic seas and NE Atlantic Ocean the LGM was characterized by seasonal open water, iceberg melting, and relatively warm summer SST. Atlantic surface water was continuously present [e.g., Sarnthein *et al.*, 1995; Weinelt *et al.*, 1996; Rasmussen *et al.*, 1996c; Pflaumann *et al.*, 2003]. Small specimens of *T. quinqueloba* were abundant in the Iceland Sea at 69°N [Kandiano and Bauch, 2002], at the Svalbard margin, and in the Fram Strait indicating that the Atlantic inflow reached to >76°N [e.g., Hebbeln *et al.*, 1994; Knies *et al.*, 1999; Hald *et al.*, 2001; Nørgaard-Pedersen *et al.*, 2003; Rasmussen *et al.*, 2007].

[29] The results from LINK17, ENAM93-21, and ENAM33 show that the warm Atlantic surface water pushed northward into the Nordic seas both east and west of the Faeroe Islands (Figures 4, 5, and 6). Overall, the temperatures must have been lowest at ENAM93-21. The high number of foraminifera, the high percentages of *T. quinqueloba*, and the high concentration of IRD point to open water during the summer and a high-nutrient level. The conditions were probably not very different from the modern environment in the central Nordic seas in areas with Arctic surface water and summer SST from 3 to 5°C [Johannessen *et al.*, 1994].

[30] However, the conditions were clearly very unstable, as indicated by the high-amplitude fluctuations that characterize most of the measured parameters, in particular the percentages of *T. quinqueloba* and the concentrations of IRD and planktonic foraminifera. The fluctuations, which are most pronounced in the LINK17 and ENAM33 cores, indicate a great instability in the inflow of Atlantic water. Similar unstable conditions have been observed in the DAPC-02 core from south of the Greenland-Scotland Ridge [Knutz *et al.*, 2007] (Figure 8). They are also in agreement with observations from the Fennoscandian and British ice sheets, which experienced several rapid fluctuations during the LGM [Dahlgren and Vorren, 2003; Hjelstuen *et al.*, 2004; McCabe *et al.*, 2007].

[31] In LINK17 and ENAM33, the sea surface temperatures seem to have been dropping during the LGM as indicated by a decrease in relative abundance of *T. quinqueloba* (Figures 4 and 6). At the end of the LGM the temperatures were probably fairly similar north and south of the ridge (Figure 6).

6.2.4. Regional Events During the LGM-HE1 Transition

[32] From about 21,500 calibrated years B.P. and into the early Holocene the climate is strongly influenced by a steep increase in insolation, affecting both the LGM-HE1 transition and the following deglaciation period [e.g., Berger, 1978]. However, comparison of cores from the Atlantic Ocean and Nordic seas shows that the response of the ocean differed significantly from area to area, depending mainly on the supply of meltwater and the distance to the nearest ice sheets. In core 23415, located centrally in the northeast-

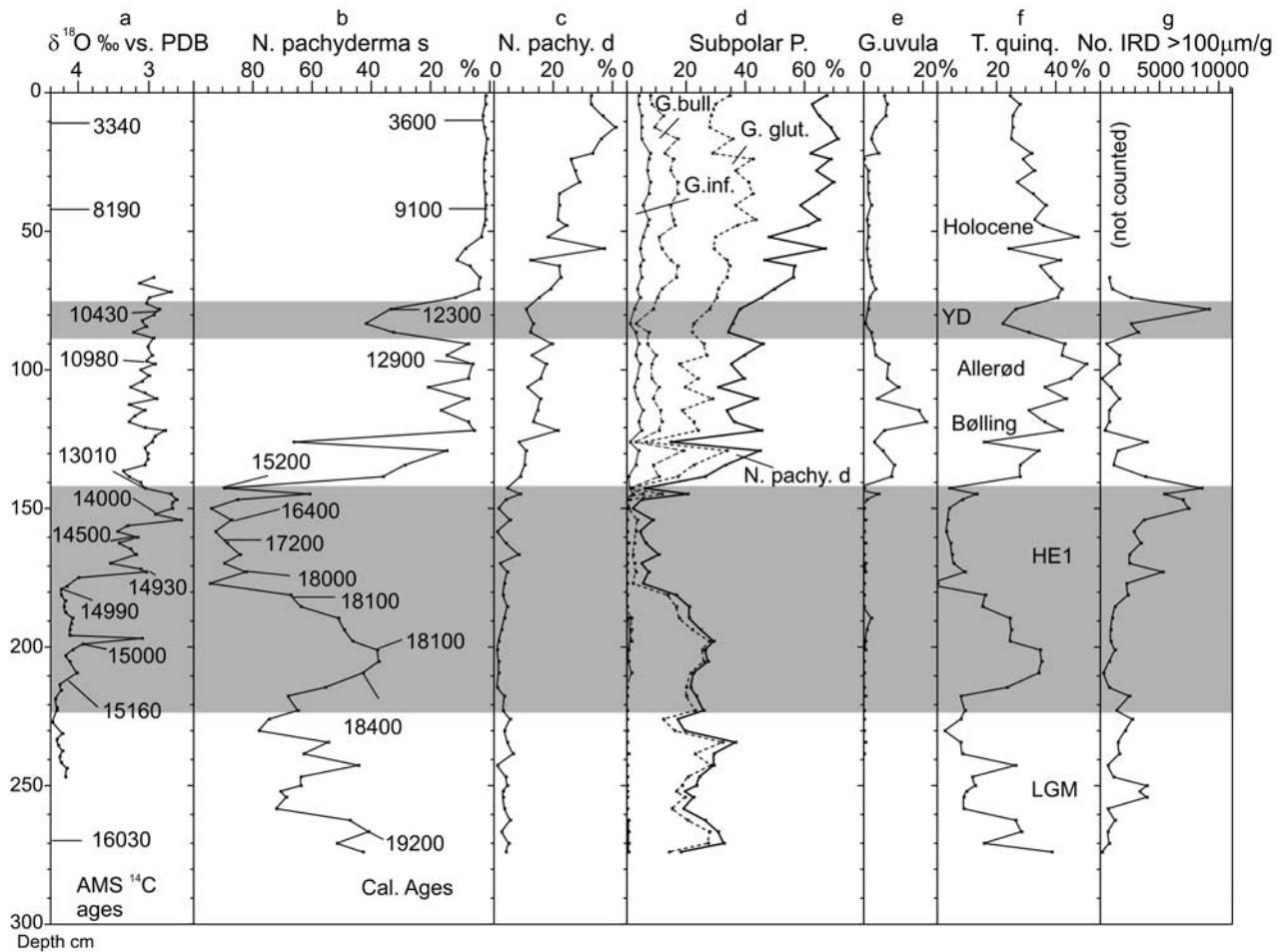


Figure 8. Planktic $\delta^{18}\text{O}$ values and variation in relative abundance of foraminifera species for core DAPC-02. (a) Planktic $\delta^{18}\text{O}$ values measured on *N. pachyderma s* with AMS ^{14}C dates marked. (b) Percent of *N. pachyderma s* with calibrated ages marked. (c) Percent of *N. pachyderma d*. (d) Cumulative percent of subpolar species *G. inflata*, *G. bulloides*, *G. glutinata*, and *N. pachyderma d*. (e) Percent of *G. uvula*, and (f) percent *T. quinqueloba*. (g) Number of IRD grains $>100\ \mu\text{m}$ per gram dry weight sediment. Abbreviations as in Figures 3 and 4.

ern Atlantic Ocean at 53°N , 19°W (Figure 1), the surface water temperatures started to increase from about 19,000 calibrated years B.P. and the warming continued throughout HE1 [Weinelt *et al.*, 2003]. In ENAM33, from 61°N , 11°W , a gradual increase in the percentage of subpolar species indicates a fairly similar pattern, although the temperatures here were considerably lower than at the 23415 core (Figures 1 and 6). At a tropical site near Grenada in the Caribbean the warming began somewhat earlier at about 19,200 calibrated years B.P., but also here temperatures remained high throughout HE1 [Rühlemann *et al.*, 1999, 2004].

[33] The cores discussed above reveal a continuous temperature increase throughout HE1. However, most cores from the northeastern Atlantic Ocean and Nordic seas show a trend that is nearly the opposite. For example, in DAPC-02, taken at 59°N in the northern Rockall Trough the surface temperatures started to increase from about 18,500 calibrated years B.P. [Knutz *et al.*, 2002] (Figure 8).

However, the increase lasted only a few hundred years. From 18,100 calibrated years B.P. the temperatures dropped abruptly, as indicated by an increase in the percentage of *N. pachyderma s* from ~ 40 to 90% , and they remained low for the rest of HE1 until the beginning of the Bølling interstadial at 15,300 calibrated years B.P. A fairly similar pattern is seen in core V29-191 from 54°N , 16°W [Lagerklint and Wright, 1999] (Figure 1) and further north in the southern Norwegian Sea in cores ENAM93-21 and LINK17 (Figure 4). In DAPC-02 the maximum percentage of *N. pachyderma s* during the early warm phase was 40% as compared to 70% in LINK17 and $75\text{--}80\%$ in ENAM93-21, indicating a clear south to north temperature decrease.

[34] We attribute the temperature drop in the southern Norwegian Sea and north-easternmost Atlantic Ocean from about 18,100 calibrated years B.P. to the impact of icebergs and meltwater. The strong orbital forcing together with the northward penetration of warm water and an increase in the thermohaline circulation [e.g., Rasmussen *et al.*, 1996b]

probably caused a rapid breakdown of the Fennoscandian and British ice sheets delivering huge amounts of icebergs and cold fresh water to the Nordic seas and eastern Atlantic ocean [e.g., McIntyre and Molino, 1996; Knutz et al., 2002, 2007; Leksens et al., 2005]. The meltwater plume spread southward into the northeastern Atlantic as indicated by the ^{14}C dates of the temperature drop in the various cores. A few cores, such as 23415 and ENAM33 placed centrally in the North Atlantic Drift, were only lightly affected by meltwater or were reached only during short-lasting excursions. Together with similar contributions from the Laurentide ice sheet, the meltwater plume, and icebergs laid the foundation for the stratified and extremely cold North Atlantic HE1-ocean.

6.2.5. Bølling and Allerød Interstadials

[35] The increase in the relative abundance of *N. pachyderma* during the Bølling and Allerød interstadials in all records including DAPC-02 indicates higher temperatures and a probably stronger inflow of Atlantic surface water as compared with the earlier interstadials (Figures 4, 5, 6, and 7, Table 2). Over the entire eastern North Atlantic region the transition to the Bølling interstadial at approximately 15,300 calibrated years B.P. is characterized by a large and abrupt surface water warming [e.g., Koç et al., 1996; Austin and Kroon, 1996; Kroon et al., 1997; Lagerklint and Wright, 1999; Rasmussen et al., 2002, 2003a]. This is also apparent in the ENAM33 and DAPC-02 cores from south of the Greenland-Scotland Ridge (Figures 6 and 7). On the Norwegian shelf, the surface water experiences a similar abrupt temperature increase [Lehman and Keigwin, 1992; Klitgaard-Kristensen et al., 1998].

[36] However, the records from the open ocean settings north of the Greenland-Scotland Ridge show only a minor surface warming [Sejrup et al., 1984; Koç and Jansen, 1992; Rasmussen et al., 1996a, 2002; Hald and Aspel, 1997]. Furthermore, in LINK17 and ENAM93-21 the warming is stepwise and not abrupt (Figures 4 and 5). In both cores the early warming is followed by a marked cooling centered on 14,500 calibrated years B.P. and lasting about 100 years, which corresponds in time to the Older Dryas event [e.g., Koç and Jansen, 1992; Björck et al., 1996; Friedrich et al., 2001]. The Bølling-Allerød period has the highest rate of sea level rise and the highest runoff of meltwater during the deglaciation [Fairbanks, 1989]. The rate peaked during the Older Dryas [Stanford et al., 2006].

[37] A characteristic planktonic species for the Bølling and Allerød interval in LINK17 is *G. uvula*. It has a similar distribution pattern in ENAM33 and DAPC-02 (Figures 6 and 7). The species is indicative of a high-food supply demonstrating that the Bølling and Allerød interstadials were productive periods. This high-food level may be related to the high runoff.

7. Conclusions

[38] LINK17 is located north of Scotland directly below the warmest inflow of Atlantic water to the Norwegian seas. The core contains a high-resolution record of planktonic foraminifera from the coldest part of the last glacial period

and succeeding deglaciation (34–10 calibrated ka). It comprises Heinrich events HE3, HE2, and HE1; interstadials IS4, IS3, IS2, and IS1; and the LGM. LINK17 contains a more diverse planktonic fauna than any of the hitherto investigated cores from the Norwegian Sea. For the first time, we have direct foraminiferal evidence for variations in the inflow of Atlantic surface water during Dansgaard-Oeschger events.

[39] During Heinrich events HE3, HE2, and HE1 and stadial 4 the conditions at LINK17 were polar with a stratified water column and surface water temperatures lower than 2°C. There was no surface inflow of Atlantic water to the Nordic seas. During interstadials IS4, IS3, and IS2, the sea surface temperatures at LINK17 rose to 4–7°C, almost as high as in the North Atlantic. Atlantic surface water flowed into the Nordic seas along the British and Norwegian shelves. Conditions west of the Faeroe-Shetland Channel were polar with temperatures not exceeding 4°C and a reduced inflow of colder Atlantic surface water.

[40] The climate shifts at the beginning and end of the interstadials were extremely rapid. The warmings took between 10 and 30 years. Maximum warmth lasted generally about 50–100 years. The coolings were slightly slower taking normally from 100 to 200 years. In LINK17 the Bølling and Allerød interstadials differ from the glacial interstadials by more gradual warmings and coolings and by strongly fluctuating surface temperatures, probably as a result of variations in the runoff from the melting continental ice sheets.

[41] Heinrich events HE3 and HE2 were preceded by long-time intervals with slowly increasing surface water temperatures. The intervals, termed interstadial F-IS5b and interstadial F-IS3b, lasted between 1600 and 2000 years, corresponding to one or two “normal” Dansgaard-Oeschger events. The time equivalent intervals in the Greenland ice cores are monotonous with $\delta^{18}\text{O}$ values intermediate between interstadial and stadial values. During the two intervals, conditions at LINK17 were interstadial-like probably with deepwater formation, whereas those in the central and western part were more polar. After 1600 to 2000 years of slowly increasing surface water temperatures the conditions suddenly shifted from interstadial to stadial. The shifts from F-IS5b to HE3 and from F-IS3b to HE2 took probably only 10–20 years. We suggest that the temperature increase during the interstadials destabilized the marine ice shelves resulting in a sudden release of icebergs and meltwater to the sea. Reaching a threshold the ocean became stratified and the convection stopped.

[42] In LINK17 the LGM was almost as warm as the interstadials. Contrary to the interstadials, inflow of warm Atlantic surface water occurred on both sides of the Faeroe Islands. The conditions were unstable with several shifts in surface water temperatures.

[43] **Acknowledgments.** The R/V *DANA* cruise August 2000 was funded by the SNF (Danish Natural Science Research Foundation) as part of the LINK project. The study of LINK17 was financed by UNIS as a part of the ESF-EuroClimate Program Resolution (grant 04-ECLIM-FP33). We warmly thank T. Nielsen, A. Kuijpers, and J. Boserup (GEUS) for invaluable help in the coring procedures. Peter Konradi is thanked for his help in the post cruise core handling and I. S. Nielsen (GEUS) for the core logging.

References

- Austin, W. E. N., and D. Kroon (1996), Late glacial sedimentology, foraminifera, and stable isotope stratigraphy of the Hebridean continental shelf, northwest Scotland, *Geol. Soc. Spec. Publ.*, 111, 187–213.
- Bé, A. W. H. and D. S. Tolderlund (1971), Distribution and ecology of living planktonic foraminifera in surface waters of the Atlantic and Indian Oceans, in *The Micropaleontology of the Oceans*, edited by B. M. Funnell and W. R. Riedel, pp. 105–149, Cambridge Univ. Press, New York.
- Berger, A. L. (1978), Long-term variations of caloric insolation resulting from Earth's orbital elements, *Quat. Res.*, 9, 139–167.
- Björck, S., et al. (1996), Synchronized terrestrial-atmospheric deglacial records around the North Atlantic, *Science*, 274, 1155–1160.
- Boltovskoy, E., D. Boltovskoy, N. Correa, and F. Brandini (1996), Planktic foraminifera from the southwestern Atlantic (30°–60°S): Species-specific patterns in the upper 50 m, *Mar. Micropaleontol.*, 28, 53–72.
- Bond, G., W. S. Broecker, S. J. Johnsen, J. McManus, L. Labeyrie, J. Jouzel, and G. Bonani (1993), Correlations between climate records from North Atlantic sediments and Greenland ice, *Nature*, 365, 143–147.
- Carstens, J., D. Hebbeln, and G. Wefer (1997), Distribution of planktic foraminifera at the ice margin in the Arctic (Fram Strait), *Mar. Micropaleontol.*, 29, 257–269.
- Dahlgren, K. I. T., and T. O. Vorren (2003), Sedimentary environment and glacial history during the last 40 ka of the Vøring continental margin, mid-Norway, *Mar. Geol.*, 193, 93–127.
- Dansgaard, W., et al. (1993), Evidence for general instability of past climate from a 250-kyr ice-core record, *Nature*, 364, 218–220.
- Davies, S. M., S. Wastegård, T. L. Rasmussen, S. J. Johnsen, J. P. Steffensen, K. K. Andersen, and A. Svensson (2007), Identification of the Fugloyarbanki tephra in the NGRIP ice-core: A key tie-point for marine and ice-core sequences during the last glacial period, *J. Quat. Sci.*, in press.
- Fairbanks, R. G. (1989), A 17,000-years glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation, *Nature*, 342, 637–742.
- Fairbanks, R. G., R. A. Mortlock, T.-C. Chiu, L. Cao, A. Kaplan, T. P. Guilderson, T. W. Fairbanks, A. L. Bloom, P. M. Grootes, and M.-J. Nadeau (2005), Radiocarbon calibration curve spanning 0–50,000 years B.P. based on paired $^{230}\text{Th}/^{238}\text{U}$ and ^{14}C dates on pristine corals, *Quat. Sci. Rev.*, 24, 1781–1796.
- Friedrich, M., B. Kromer, K. F. Kaiser, M. Spurk, K. A. Hughen, and S. J. Johnsen (2001), High-resolution climate signals in the Bølling-Allerød interstadial (Greenland Interstadial 1) as reflected in European tree-ring chronologies compared to marine varves and ice-core records, *Quat. Sci. Rev.*, 20, 1223–1232.
- Fronval, T., E. Jansen, J. Bloemendal, and S. J. Johnsen (1995), Oceanic evidence for coherent fluctuations in Fennoscandian and Laurentide ice sheets on millennium timescales, *Nature*, 374, 443–446.
- Hald, M., and R. Aspeli (1997), Rapid climatic shifts of the northern Norwegian Sea during the last deglaciation and the Holocene, *Boreas*, 26, 15–28.
- Hald, M., T. Dokken, and G. Mikalsen (2001), Abrupt climatic change during the last interglacial-glacial cycle in the polar North Atlantic, *Mar. Geol.*, 176, 121–137.
- Hansen, B., and S. Østerhus (2000), North Atlantic-Norwegian Sea exchanges, *Prog. Oceanogr.*, 45, 109–208.
- Hebbeln, D., T. Dokken, E. S. Andersen, M. Hald, and A. Elverhoi (1994), Moisture supply for northern ice-sheet growth during the Last Glacial Maximum, *Nature*, 370, 357–360.
- Heinrich, H. (1988), Origin and consequences of cyclic ice rafting in the northeast Atlantic Ocean during the past 130,000 years, *Quat. Res.*, 29, 142–152.
- Hjelstuen, B. O., H.-P. Sejrup, H. Haflidason, A. Nygård, M. Berstad, and G. Knorr (2004), Late Quaternary seismic stratigraphy and geological development of the south Vøring margin, Norwegian Sea, Norwegian Sea, *Quat. Sci. Rev.*, 23, 1847–1865.
- Hulbe, C. L., D. R. MacAyeal, G. H. Denton, J. Klemm, and T. V. Lowell (2004), Catastrophic ice shelf breakup as the source of Heinrich event icebergs, *Paleoceanography*, 19, PA1004, doi:10.1029/2003PA000890.
- Johannessen, T., E. Jansen, A. Flato, and A. C. Ravelo (1994), The relationship between surface water masses, oceanographic fronts and paleoclimatic proxies in surface sediments of the Greenland, Iceland, Norwegian seas, in *Carbon Cycling in the Glacial Ocean: Constraints on the Ocean's Role in Global Change*, NATO ASI Ser. I, vol. 117, edited by R. Zahn et al., pp. 61–86, Springer, Berlin.
- Johnsen, S. J., H. B. Clausen, W. Dansgaard, K. Fuhrer, N. Gundestrup, C. U. Hammer, P. Iversen, J. Jouzel, B. Stauffer, and J. P. Steffensen (1992), Irregular glacial interstadials recorded in a new Greenland ice core, *Nature*, 359, 311–313.
- Kandiano, E. S., and H. A. Bauch (2002), Implications of planktic foraminiferal size fractions for the glacial-interglacial paleoceanography of the polar North Atlantic, *J. Foraminiferal Res.*, 32, 245–251.
- Kemle-von Mücke, S. and C. Hemleben (1999), Foraminifera, in *South Atlantic Zooplankton*, edited by D. Boltovskoy, pp. 43–73, Backhuys, Leiden, Netherlands.
- King, E. L., H. P. Sejrup, H. Haflidason, A. Elverhoi, and I. Aarseth (1996), Quaternary seismic stratigraphy of the North Sea Fan: Glacially-fed gravity flow aprons, hemipelagic sediments, and large submarine slides, *Mar. Geol.*, 130, 293–315.
- Klitgaard-Kristensen, D., T. L. Rasmussen, H.-P. Sejrup, H. Haflidason, and T. C. E. van Weering (1998), Rapid changes in the oceanic fronts in the Norwegian Sea during the last deglaciation: Implications for the Younger Dryas cooling event, *Mar. Geol.*, 152, 177–188.
- Knies, J., C. Vogt, and R. Stein (1999), Late Quaternary growth and decay of the Svalbard/Barents Sea ice sheet and paleoceanographic evolution in the adjacent Arctic Ocean, *Geo Mar. Lett.*, 18, 195–202.
- Knutz, P. C., I. R. Hall, R. Zahn, T. L. Rasmussen, A. Kuijpers, M. Moros, and N. J. Shackleton (2002), Multidecadal ocean variability and NW European ice sheet surges during the last deglaciation, *Geochem. Geophys. Geosyst.*, 3(12), 1077, doi:10.1029/2002GC000351.
- Knutz, P. C., R. Zahn, and I. R. Hall (2007), Centennial-scale variability of the British Ice Sheet: Implications for climate forcing and Atlantic meridional overturning circulation during the last deglaciation, *Paleoceanography*, 22, PA1207, doi:10.1029/2006PA001298.
- Koç, N., and E. Jansen (1992), A high-resolution diatom record of the last deglaciation from the SE Norwegian Sea: Documentation of rapid climate changes, *Paleoceanography*, 7, 499–520.
- Koç, N., E. Jansen, M. Hald, and L. Labeyrie (1996), Late-glacial-Holocene sea surface temperatures and gradients between the North Atlantic and the Norwegian Sea: Implications for the Nordic heat pump, *Geol. Soc. Spec. Publ.*, 111, 177–185.
- Kroon, D., W. E. N. Austin, M. R. Chapman, and G. M. Ganssen (1997), Deglacial surface circulation changes in the northeastern Atlantic: Temperature and salinity records off NW Scotland on a century scale, *Paleoceanography*, 12(6), 755–763.
- Kuijpers, A., S. R. Troelstra, M. Wisse, S. Heier Nielsen, and T. C. E. van Weering (1998), Norwegian Sea overflow variability and NE Atlantic surface hydrography during the past 150,000 years, *Mar. Geol.*, 12, 75–99.
- Lagerklint, I. M., and J. D. Wright (1999), Late glacial warming prior to Heinrich event 1: The influence of ice rafting and large ice sheets on the timing of initial warming, *Geology*, 27, 1099–1102.
- Lehman, S. J., and L. D. Keigwin (1992), Sudden changes in North Atlantic circulation during the last deglaciation, *Nature*, 356, 757–762.
- Lekens, W. A. H., H.-P. Sejrup, H. Haflidason, G. Ø. Petersen, B. Hjelstuen, and G. Knorr (2005), Laminated sediments preceding Heinrich event 1 in the northern North Sea and southern Norwegian Sea: Origin, processes, and regional linkage, *Mar. Geol.*, 216, 27–50.
- McCabe, A. M., P. U. Clark, D. E. Smith, and P. Dunlop (2007), A revised model for the last deglaciation of eastern Scotland, *J. Geol. Soc. London*, 164, 313–316.
- McIntyre, A., and B. Molino (1996), Forcing of Atlantic equatorial and subpolar millennial cycles by precession, *Science*, 274, 1867–1870.
- Nielsen, T., T. L. Rasmussen, S. Ceramicola, and A. Kuijpers (2000), Mapping and sampling glacial and post-glacial sediments on the Faeroe Shelf and slope areas, *GEUS Rep. 2000/69*, pp. 1–23, Geol. Surv. of Den. and Greenl., Copenhagen.
- Nørgaard-Pedersen, N., R. F. Spielhagen, H. Erlenkeuser, P. M. Grootes, J. Heinemeier, and J. Knies (2003), Arctic Ocean during the Last Glacial Maximum: Atlantic and polar domains of surface water mass distribution and ice cover, *Paleoceanography*, 18(3), 1063, doi:10.1029/2002PA000781.
- North Greenland Ice Core Project members (2004), High-resolution record of Northern Hemisphere climate extending into the last interglacial period, *Nature*, 431, 147–151.
- Oppo, D. W., and S. J. Lehman (1995), Suborbital timescale variability of North Atlantic Deep Water during the past 200,000 years, *Paleoceanography*, 10(5), 901–910.
- Orvik, K. A., and P. Niiler (2002), Major pathways of Atlantic water in the northern North Atlantic and Nordic Seas toward Arctic, *Geophys. Res. Lett.*, 29(19), 1896, doi:10.1029/2002GL015002.
- Ostermann, D. R., and W. D. Curry (2000), Calibration of stable isotope data: An enriched $\delta^{18}\text{O}$ standard used for source gas mixing detection and correction, *Paleoceanography*, 15, 353–360.

- Ottens, J. J., and A. J. Nederbragt (1992), Planktic foraminifera diversity as indicator of ocean environments, *Mar. Micropaleontol.*, **19**, 13–28.
- Parker, F. L. (1962), Planktonic foraminiferal species in Pacific sediments, *Micropaleontology*, **8**, 219–254.
- Pflaumann, U., J. Duprat, C. Pujol, and L. D. Labeyrie (1996), SIMMAX: A modern analog technique to deduce Atlantic sea surface temperatures from planktonic foraminifera in deep-sea sediments, *Paleoceanography*, **11**, 15–35.
- Pflaumann, U., et al. (2003), Glacial North Atlantic: Sea-surface conditions reconstructed by GLAMAP 2000, *Paleoceanography*, **18**(3), 1065, doi:10.1029/2002PA000774.
- Rasmussen, T. L., and E. Thomsen (2004), The role of the North Atlantic Drift in the millennial timescale glacial climate fluctuations, *Paleogeogr. Palaeoclim. Palaeoecol.*, **210**, 101–116.
- Rasmussen, T. L., E. Thomsen, T. C. E. van Weering, and L. Labeyrie (1996a), Rapid changes in surface and deep water conditions at the Faeroe Margin during the last 58,000 years, *Paleoceanography*, **11**(6), 757–772.
- Rasmussen, T. L., E. Thomsen, L. Labeyrie, and T. C. E. van Weering (1996b), Circulation changes in the Faeroe-Shetland Channel correlating with cold events during the last glacial period (58–10 ka), *Geology*, **24**, 937–940.
- Rasmussen, T. L., T. C. E. van Weering, and L. Labeyrie (1996c), High resolution stratigraphy of the Faeroe-Shetland Channel and its relation to North Atlantic paleoceanography: The last 87 kyr, *Mar. Geol.*, **131**, 75–88.
- Rasmussen, T. L., E. Thomsen, and T. C. E. van Weering (1998), Cyclic sedimentation on the Faeroe Drift 53–10 ka BP related to climatic variations, *Geol. Soc. Spec. Publ.*, **129**, 255–267.
- Rasmussen, T. L., D. Bäckström, J. Heinemeier, D. Klitgaard-Kristensen, P. C. Knutz, A. Kuijpers, S. Lassen, E. Thomsen, S. R. Troelstra, and T. C. E. van Weering (2002), The Faeroe-Shetland Gateway: Late Quaternary water mass exchange between the Nordic seas and the northeastern Atlantic, *Mar. Geol.*, **188**, 165–192.
- Rasmussen, T. L., E. Thomsen, S. R. Troelstra, A. Kuijpers, and M. A. Prins (2003a), Millennial-scale glacial variability versus Holocene stability: Changes in planktic and benthic foraminifera faunas and ocean circulation in the North Atlantic during the last 60,000 years, *Mar. Micropaleontol.*, **47**, 143–176.
- Rasmussen, T. L., S. Wastegård, A. Kuijpers, T. C. E. van Weering, J. Heinemeier, and E. Thomsen (2003b), Stratigraphy and distribution of tephra layers in marine sediment cores from the Faeroe Islands, North Atlantic, *Mar. Geol.*, **199**, 263–277.
- Rasmussen, T. L., E. Thomsen, M. A. Ślubowska, S. Jessen, A. Solheim, and N. Koç (2007), Paleocceanographic evolution of the SW Svalbard margin (76°N) since 20,000 ¹⁴C yr BP, *Quat. Res.*, **67**, 100–114.
- Rühlemann, C., S. Mulitz, P. J. Müller, G. Wefer, and R. Zahn (1999), Warming of the tropical Atlantic Ocean and slowdown of thermohaline circulation during the last deglaciation, *Nature*, **402**, 511–514.
- Rühlemann, C., S. Mulitz, G. Lohmann, A. Paul, M. Prange, and G. Wefer (2004), Intermediate depth warming in the tropical Atlantic related to weakened thermohaline circulation: Combining paleoclimate data and modeling results for the last deglaciation, *Paleoceanography*, **19**, PA1025, doi:10.1029/2003PA000948.
- Sarnthein, M., et al. (1995), Variations in Atlantic surface ocean paleoceanography, 50°–80°N: A time-slice record of the last 30,000 years, *Paleoceanography*, **10**, 1063–1094.
- Sejrup, H.-P., E. Jansen, H. Erlenkeuser, and H. Høltedahl (1984), New faunal and isotopic evidence on the Late Weichselian-Holocene oceanographic changes in the Norwegian Sea, *Quat. Res.*, **21**, 74–84.
- Sejrup, H.-P., E. King, I. Aarseth, H. Haflidason, and A. Elverhøy (1996), Quaternary erosion and depositional processes: Western Norwegian fjords, Norwegian Channel, and the North Sea Fan, *Geol. Soc. London*, **117**, 187–202.
- Sherwin, T. J., W. R. Turrell, D. R. G. Jeans, and S. Dye (1999), Eddies and a mesoscale deflection of the slope current in the Faeroe-Shetland Channel, *Deep Sea Res.*, **146**, 415–438.
- Stanford, J. D., E. J. Rohling, S. E. Hunter, A. P. Roberts, S. O. Rasmussen, E. Bard, J. McManus, and R. G. Fairbanks (2006), Timing of meltwater pulse 1a and climate responses to meltwater injections, *Paleoceanography*, **21**, PA4103, doi:10.1029/2006PA001340.
- van Aken, H. M. (1988), Transports of water masses through the Faeroese Channels determined by an inverse method, *Deep Sea Res.*, **35**, 595–617.
- van Kreveld, S., M. Sarnthein, H. Erlenkeuser, P. Grootes, S. Jung, M. J. Nadeau, U. Pflaumann, and A. Voelker (2000), Potential links between surging ice sheets, circulation changes, and the Dansgaard-Oeschger cycles in the Irminger Sea, 60–18 kyr, *Paleoceanography*, **15**(4), 425–442.
- Voelker, A. H. L., M. Sarnthein, P. M. Grootes, H. Erlenkeuser, C. Laj, A. Mazaud, M. J. Nadeau, and M. Schleicher (1998), The correlation of marine ¹⁴C ages from the Nordic seas with the GISP2 isotope record: Implications for ¹⁴C calibration beyond 25 ka B. P., *Radio-carbon*, **40**, 517–534.
- Wastegård, S., T. L. Rasmussen, A. Kuijpers, T. Nielsen, and T. C. E. van Weering (2006), Composition and origin of ash zones from Marine Isotope Stages 3 and 2 in the North Atlantic, *Quat. Sci. Rev.*, **25**, 2409–2419.
- Weinelt, M., M. Sarnthein, U. Pflaumann, H. Schulz, S. Jung, and H. Erlenkeuser (1996), Ice-free Nordic seas during the Last Glacial Maximum? Potential sites of deepwater formation, *Paleoclimates*, **1**, 283–309.
- Weinelt, M., E. Vogelsang, M. Kucera, U. Pflaumann, M. Sarnthein, A. Voelker, H. Erlenkeuser, and B. A. Malmgren (2003), Variability of North Atlantic heat transfer during MIS 2, *Paleoceanography*, **18**(3), 1071, doi:10.1029/2002PA000772.

T. L. Rasmussen, Woods Hole Oceanographic Institute, 266 Woods Hole Road, Woods Hole, MA 02543, USA. (tine.rasmussen@ig.uit.no)

E. Thomsen, Department of Earth Sciences, University of Aarhus, DK-8000 Aarhus, Denmark.